Geodynamic evolution of southern Costa Rica related to low-angle subduction of the Cocos Ridge: constraints from thermochronology

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Abstract

The Late Tertiary shallow subduction of the Cocos ridge under the Caribbean plate controlled the evolution of the Cordillera de Talamanca in southeast Costa Rica, which is a mountain range that consists mainly of granitoids formed in a volcanic arc setting. Fission track thermochronology using zircon and apatite, as well as 40Ar–39Ar and Rb–Sr age data of amphibole and biotite in granitoid rocks constrain the thermal history of the Cordillera de Talamanca and the age of onset of subduction of the Cocos ridge. Shallow intrusion of granitoid melts resulted in fast and isobaric cooling. A weighted mean zircon fission track age (13 analyses) and Rb–Sr biotite ages of about 10 Ma suggest rapid cooling and give minimum ages for granitoid emplacement. In some cases 40Ar–39Ar and Rb–Sr apparent ages of amphibole and biotite are younger than the zircon fission track ages, which can be attributed to partial resetting by hydrothermal alteration. Apatite fission track ages range from 4.8 to 1.7 Ma but show no correlation with the 3090-m elevation span over which they were sampled. The apatite ages seem to indicate rapid exhumation caused by tectonic and isostatic processes. The combination of the apatite fission track ages with subduction parameters of the Cocos plate such as subduction angle, plate convergence rate and distance of the Cordillera de Talamanca to the trench implies that the Cocos ridge entered the Middle America Trench between 5.5 and 3.5 Ma. © 2002 Elsevier Science B.V. All rights reserved.

Keywords: Ridge subduction; Fission track dating; Geochronology; Thermal history; Cocos Ridge; Costa Rica

1. Introduction

The shallow subduction of oceanic ridges below continental margins has profound effects on the geodynamic evolution of the upper plate. Subduction of oceanic ridges in the east Pacific realm occurs off Ecuador (Carnegie ridge, Gutscher et al., 1999), northern Chile (Nazca ridge, Barazangi and Isacks, 1976; Pilger, 1981) and southern Costa Rica (Cocos ridge, Kolarsky et al., 1995). Gravity, seismic refraction surveys and isostatic modeling have shown that oceanic ridges are characterized by thickened crust and high heat flow (Detrick and Watts, 1979; Kogan, 1979). The resulting buoyancy of particularly young ridges leads to low-angle subduction of oceanic crust and increased coupling between the lower and upper plates (Cross and Pilger, 1982; Behrmann et al., 1994; Gutscher et al., 1999). Displacement of the mantle wedge as a consequence of low-angle subduction may result in termination of arc volcanism as observed in
Fig. 1. (A) Plate configuration of Central and South America (after Gutscher et al., 1999). Inlay refers to map B. (B) Major geologic units, tectonic features and sedimentary basins in Costa Rica. The position of the Cocos ridge is indicated by the 1000-m bathymetric line. Inlay shows the location of the study area with the Chirripó profile from the town of San Gerardo to Mount Chirripó, and the Buenos Aires profile, near the town of Ujarraz.
northern Chile and southern Costa Rica (McGeary et al., 1985; de Boer et al., 1995; Kolarsky et al., 1995).

Subduction of the young aseismic Cocos ridge at the southern Middle America Trench (Fig. 1) has produced a number of distinct geodynamic characteristics. High-resolution seismicity data reveal that the subduction angle is about 60° in central Costa Rica. In southern Costa Rica, which is the location of ridge subduction and the target area of the present study, the data reveal an angle of 30° for the first 60 km of the subducting plate, thereafter the plate is assumed to continue almost horizontal (Adamek et al., 1987; Protti et al., 1995). Paleobathymetric and stratigraphic data document the Early to Late Pliocene emergence of fore-arc and back-arc basins in southern Costa Rica due to isostatic reequilibration induced by ridge subduction (Collins et al., 1995). This is manifested by a change from marine to continental sedimentation in the outer arc of the Burica and Osa Peninsulas and the Terraba fore-arc basin (Fig. 1). Chronostratigraphy of sediments from the back-arc Limón basin (Fig. 1) indicates Pleistocene uplift rates of 0.05 mm/year, attributed to the subduction of the Cocos ridge (McNeill et al., 2000). Uplift rates based on the age and paleobathymetry of marine microfauna in post Miocene shelf and slope deposits are 0.2 mm/year at the Osa Peninsula and 0.94 mm/year at the Burica Peninsula (Fig. 1; Corrigan et al., 1990).

The relationship between upper plate processes and ridge subduction has been discussed by several authors (Barazangi and Isacks, 1976; Pilger, 1981; Gutscher et al., 1999). In southern Costa Rica, the tectonic inversion of fore-arc and back-arc basins has been attributed to horizontal contraction caused by increased coupling between the oceanic and continental plates as a response to low-angle subduction of the Cocos ridge (Kolarsky et al., 1995). Surface uplift of the magmatic arc above the subducted ridge in the Cordillera de Talamanca may have been accomplished isostatically as a result of horizontal shortening due to underthrusting of the Cocos ridge (Fig. 1; Drummond et al., 1995; Kolarsky et al., 1995). Alternatively, Abratis and Wörner (2001) suggested that thermal processes due to slab-window formation contributed to uplift in the Cordillera de Talamanca. Ridge subduction in southern Costa Rica probably led to: (1) termination of calc-alkaline igneous activity (Drummond et al., 1995), (2) erosion of the volcanic rocks over an arc-parallel distance of about 180 km (Fig. 1; Ballmann, 1976; de Boer et al., 1995), and (3) emplacement of adakitic melts apparently derived from melting of the subducted portion of the Cocos ridge in the Cordillera de Talamanca beginning at 3.6 Ma, (Defant et al., 1991a,b, 1992; de Boer et al., 1995; Abratis and Wörner, 2001).

However, the precise correlation between the upper plate processes and the Cocos ridge subduction remains unclear, mainly due to a lack of geochronological data directly dating uplift. In order to constrain the time of incipient ridge subduction, age data on the magmatic activity, and the precise cooling and exhumation history of the granitoid rocks of the upper plate are needed.

The age of incipient subduction of the Cocos ridge under the southern Costa Rican magmatic arc has been debated over the last 20 years. Collins et al. (1995) and Meschede et al. (1998) postulated that subduction of the Cocos ridge initiated at ca. 3.6 Ma and between 4 and 3 Ma, respectively. Kolarsky et al. (1995) suggested an age of 5 Ma based on the development of thrust faults in the Terraba basin (Fig. 1). These results clearly differ from those of earlier studies by Lonsdale and Klitgord (1978), Corrigan et al. (1990), and Gardner et al. (1992), who argued that the ridge subduction occurred fairly recently, between 0.5 and 1 Ma. The purpose of this paper is to date the beginning of Cocos ridge subduction as well as to quantify the effects caused by its buoyancy on the respective section of the Middle American Arc, the Cordillera de Talamanca. Fission track thermochronology and 40Ar–39Ar and Rb–Sr mineral ages from granitoid plutons of the Cordillera de Talamanca are employed to unravel the subsequent cooling history of these rocks.

2. Geological background

The Quaternary volcanic arc of Central America extends over a length of about 1000 km from central Mexico to central Costa Rica and comprises about 40 active or dormant volcanoes. Average spacing between the major stratovolcanoes is about 26 km, which makes the Middle America Volcanic Arc one of the volcanically most active convergent plate margins on Earth (Carr and Stoiber, 1990; de Boer et al., 1995). A volcanic gap of about 180 km in this
arc extends between the active Irazú–Turrialba complex in central Costa Rica and the historically active Barú complex further southeast in the Cordillera de Panamá Occidental (Fig. 1). The section of eroded volcanic edifices roughly coincides with the width of the subducting Cocos ridge (Fig. 1). The northeast-trending Cocos ridge is about 200–300 km wide and consists of thickened oceanic crust that rises 2–2.5 km above the ocean floor of the Cocos plate (von Huene et al., 1995). The ridge contains a 2-km-thick, low-density volcanic layer (Bently, 1974), which makes the Cocos ridge more buoyant than the adjacent Early to Middle Miocene oceanic crust (Hey, 1977; Meschede et al., 1998).

The Cordillera de Talamanca forms that segment of the magmatic arc with the volcanic gap above the subducting Cocos ridge. The topography of the Cordillera de Talamanca is asymmetric in cross-section whereby the southwestern flank of the Cordillera is steeper than the northeastern one. Structurally, the southwestern flank is dominated by reverse and normal faults, whereas thrust faults characterize the Caribbean side of the Cordillera de Talamanca (Kolasky et al., 1995). Based on petrographic, geochemical, and geochronological information (Bellon and Tournon, 1978; Tournon, 1984; Defant et al., 1992, Drummond et al., 1995), Drummond et al. (1995) divided the volcanic and intrusive rocks of the Cordillera de Talamanca into four lithologic units: (1) Middle Oligocene tholeiitic gabbros; (2) Middle Oligocene calc-alkaline plutonic rocks of the El Barú complex; (3) Middle to Late Miocene calc-alkaline Talamanca Intrusive Suite; (4) Plio–Pleistocene andesitic and adakitic volcanic rocks, whose geochemical characteristics are consistent with an origin from subducted oceanic crust of the Cocos ridge.

3. Sample description

Samples for fission track thermochronology, \(^{40}\text{Ar}–^{39}\text{Ar}\) and Rb–Sr age determination were collected from granitoid rocks of the central Cordillera de Talamanca along two transects, in the Chirripó National Park and near Buenos Aires, respectively (Fig. 1). Samples weighing 5–8 kg were collected at altitudes between 500 and 1200 m in the Buenos Aires area (samples 77, 78, 78A, 79, 80) and at altitudes between 1200 and 3800 m in the Chirripó area (samples 20, 22, 48, 55, 55A, 57, 58, 59). Granitoid rocks containing magmatic hornblende and biotite occur below 2000 m, while rocks above this elevation are free of biotite, and hornblende is replaced by actinolite. The chemical composition of amphibole and biotite was determined by electron microprobe analysis. The amphibole in samples 48, 78 and 80 are magnesio-hornblende (nomenclature of Deer et al., 1992) and are replaced by actinolite at grain boundaries and intragranular fractures. The amphibole in samples 55, 55A, 57, 58 and 59 are actinolite. Electron microprobe analyses and inspection of biotite in thin sections indicate chlorite–biotite intergrowths due to biotite chloritization in samples 48 and 55A (Gräfe, 1998).

4. Results

4.1. \(^{40}\text{Ar}–^{39}\text{Ar}\) age determination

\(^{40}\text{Ar}–^{39}\text{Ar}\) age spectra of four amphibole samples from the Chirripó region (Fig. 2) are internally discordant, indicating a complex mineralogy and crystallization/alteration history. Such age spectra indicate low-temperature amphibole growth in a variable chemical environment (Villa et al., 1996a,b). A multiphase composition of the K–Ar system in Ca-bearing minerals such as amphibole can be monitored by Ca/K ratios, which can be calculated from the Ar isotope data (Fig. 2). The variability of the Ca/K ratio can be used to detect the presence of chemically different amphibole phases, because degassing of chemically different phases will lead to variations in the Ca/K spectrum. The preferential loss of K as a consequence of alteration involving actinolitization and chloritization may increase the Ca/K ratio of a sample. Step-heating experiments of all samples reveal that Ca/K ratios vary during degassing of the samples, consistent with polyphase mineral composition and possibly with loss of K (Fig. 2).

The formation of successive amphibole generations requires distinct thermodynamic and kinetic conditions. Reequilibration of magmatic assemblages in a low \(p–T\) environment is only possible in the presence of fluid. We therefore suggest that retrogressive reactions such as formation of actinolite and chlorite in the
Chirripo samples were triggered by hydrothermal fluids. Other mineral reactions related to retrograde alteration and observed in thin sections of the Chirripo area are sericitization of feldspar and chloritization of biotite. Fluid-driven alteration and neo-crystallization of amphibole will be accompanied by partial loss of Ar, resulting in partially reset ages. Therefore, the presence of several retrograde breakdown products in amphibole samples from the Chirripo area obstructs a straightforward interpretation of the Ar isotope data in terms of cooling ages. This is also obvious from the age spectrum of sample 58 (Fig. 2d) which has geologically unrealistic apparent ages of > 50 Ma in the last degassing steps that can only be explained by the presence of excess Ar. In summary, the amphibole 40Ar–39Ar age spectra are affected by complex geochemical processes. Only sample 48 displays a plateau age (Fig. 2a). Since alteration features as indicated by actinolite domains are also present in this sample the apparent age of 8.7 ± 0.2 Ma has no direct geological meaning, and can only be interpreted as a minimum cooling age.

40Ar–39Ar biotite data were obtained to better understanding cooling of the rocks around 350 °C. 40Ar–39Ar age spectra for five biotite samples from the Chirripo area show plateau ages between 8.6 and 7.5 Ma (Fig. 2a–c). In samples 48 and 55A, different sieve fractions were analyzed, but only sample 55A
yields a plateau age of 7.5 ± 0.1 Ma for both size fractions (Fig. 2c). However, loss of K from the biotite samples of the Chirripo area due to formation of secondary chlorite is evidenced by the presence of fine chlorite lamellae inside biotite and/or a chlorite rim. Therefore, the apparent ⁴⁰Ar–³⁹Ar biotite ages have to be interpreted as minimum cooling ages, in a similar way as the ⁴⁰Ar–³⁹Ar amphibole data.

4.2. Rb–Sr age determination

Five samples from both study areas were investigated by the Rb–Sr method. In order to obtain a maximum spread of the data points in the isochron diagram, apatite was analyzed in addition to the whole rock and biotite fractions, since apatite has a very low Rb/Sr ratio.

<table>
<thead>
<tr>
<th>Sample</th>
<th>[Rb]</th>
<th>[Sr]</th>
<th>⁸⁷Rb/⁸⁶Sr</th>
<th>⁸⁷Sr/⁸⁶Sr *</th>
<th>Rb–Sr age</th>
<th>⁴⁰Ar–³⁹Ar age</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chirripó Sample 22 (8.4 ± 0.1 Ma; Sr₁ = 0.70384 ± 2; MSWD = 0.1)</td>
<td></td>
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<tr>
<td>Whole rock</td>
<td>67</td>
<td>562</td>
<td>0.35</td>
<td>0.70388 ± 1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Apatite (80–125 μm)</td>
<td>ND</td>
<td>ND</td>
<td>0</td>
<td>0.70384 ± 1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Biotite (125–180 μm)</td>
<td>479</td>
<td>4.56</td>
<td>305</td>
<td>0.74034 ± 1</td>
<td>8.4 ± 0.1</td>
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</tr>
<tr>
<td>Chirripó Sample 48 (8.5 ± 0.4 Ma; Sr₁ = 0.70386 ± 66; MSWD = 6.1)</td>
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<tr>
<td>Whole rock</td>
<td>55</td>
<td>639</td>
<td>0.25</td>
<td>0.70386 ± 1</td>
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<td></td>
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<tr>
<td>Apatite (80–125 μm)</td>
<td>ND</td>
<td>ND</td>
<td>0</td>
<td>0.70383 ± 1</td>
<td></td>
<td></td>
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<tr>
<td>Biotite 1 (80–125 μm)</td>
<td>366</td>
<td>7.67</td>
<td>138</td>
<td>0.72071 ± 1</td>
<td></td>
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<tr>
<td>Biotite 2 (125–180 μm)</td>
<td>368</td>
<td>5.27</td>
<td>200</td>
<td>0.72769 ± 1</td>
<td>8.5 ± 0.4</td>
<td>8.6 ± 0.1</td>
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<tr>
<td>Amphibole</td>
<td></td>
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<td>8.7 ± 0.2</td>
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<td>Sample 55</td>
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<tr>
<td>Biotite (125–180 μm)</td>
<td></td>
<td></td>
<td></td>
<td>7.9 ± 0.1</td>
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<tr>
<td>Sample 55A</td>
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<tr>
<td>Biotite (63–125 μm)</td>
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<td></td>
<td>7.5 ± 0.1</td>
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<tr>
<td>Biotite (125–180 μm)</td>
<td></td>
<td></td>
<td></td>
<td>7.5 ± 0.1</td>
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<tr>
<td>Buenos Aires Sample 78 (10.1 ± 0.1 Ma; Sr₁ = 0.70380 ± 4)</td>
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<tr>
<td>Whole rock</td>
<td>106</td>
<td>385</td>
<td>0.797</td>
<td>0.70391 ± 1</td>
<td></td>
<td></td>
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<tr>
<td>Biotite (80–125 μm)</td>
<td>518</td>
<td>4.44</td>
<td>339</td>
<td>0.75216 ± 10</td>
<td>10.1 ± 0.1</td>
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<tr>
<td>Sample 78A (10.0 ± 0.1 Ma; Sr₁ = 0.70379 ± 4)</td>
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<tr>
<td>Whole rock</td>
<td>114</td>
<td>399</td>
<td>0.827</td>
<td>0.70391 ± 1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Biotite (80–125 μm)</td>
<td>524</td>
<td>3.48</td>
<td>439</td>
<td>0.76634 ± 8</td>
<td>10.0 ± 0.1</td>
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<tr>
<td>Sample 80 (9.9 ± 0.1 Ma; Sr₁ = 0.70379 ± 4)</td>
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<td></td>
</tr>
<tr>
<td>Whole rock</td>
<td>105</td>
<td>381</td>
<td>0.798</td>
<td>0.70391 ± 1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Biotite (80–125 μm)</td>
<td>392</td>
<td>5.10</td>
<td>223</td>
<td>0.73522 ± 14</td>
<td>9.9 ± 0.1</td>
<td></td>
</tr>
<tr>
<td>Samples 78, 78A, 80 (10.0 ± 0.1 Ma; Sr₁ = 0.70379 ± 4; MSWD = 0.8)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>10.0 ± 0.1</td>
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</tr>
</tbody>
</table>

Rb and Sr concentrations are in ppm, the ages are in Ma; ⁸⁷Rb/⁸⁶Sr ratio for apatite is assumed to be zero; the age uncertainty is reported at the 95% confidence level. 2σ input errors for the age calculations are 1% for the ⁸⁷Rb/⁸⁶Sr ratios and 0.005% for the ⁸⁷Sr/⁸⁶Sr isotope ratios. For samples with higher within-run errors, the individual error on the measurement was chosen.

wr—whole rock, ap—apatite, bt—biotite, ND—not determined.

Isochrons were calculated using the ISOPLOT data reduction program (Ludwig, 1999).

* Errors are ± 2σ mean of the within-run statistics and refer to the last digits.
Moreover, apatite may be resistant to hydrothermal alteration, even in chemical environments where the Rb–Sr system of the whole rock has been disturbed. Therefore, it may indicate the initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the system (Creaser and Gray, 1992). The closure temperature of the Rb–Sr system in biotite is commonly estimated at temperatures between 400 and 300 °C (Del Moro et al., 1982; Blanckenburg et al., 1989). However, closure temperatures are dependent on cooling rates, grain size and fluid budget (Hammouda and Cherniak, 2000; Kühn et al., 2000).

The Rb–Sr data are presented in Table 1. Two biotite samples (22, 48) from the Chirripó area give similar apparent Rb–Sr ages of about 8.4 Ma. Intergrowth between biotite and chlorite suggests that the Rb–Sr biotite ages are partially reset and should be interpreted as minimum ages in terms of cooling. The large dispersion of the 4-point regression line of sample 48 (MSWD = 6.1) also indicates that the Rb–Sr isotope system is disturbed.

Rb–Sr ages for three samples from the Buenos Aires area (Table 1) were calculated with whole rock and biotite data only as there was not sufficient apatite for analysis. Concordant ages of about 10 Ma were obtained for the three samples. In contrast to biotite samples from the Chirripó area, biotites from the Buenos Aires area show very little or no chlorite intergrowth. Although sample localities are several kilometers apart, the similar ages, identical initial Sr-isotopic compositions and the lithological homogeneity of the granitoid rocks in the sampling area warrant the calculation of a 6-point isochron including all Rb–Sr data from the Buenos Aires area. The obtained age of 10.0 ± 0.1 Ma and the low isochron dispersion (MSWD = 0.8) support the interpretation that the age is an undisturbed cooling age.

4.3. Fission track thermochronology

Zircon and apatite fission track thermochronology is an important tool for elucidating the low-temperature evolution in orogenic belts as it constrains cooling of the rocks below approximately 280–240 °C (zircon, Yamada et al., 1995; Foster et al., 1996; Brandon et al., 1998; Stöckhert et al., 1999) and 120–110 °C (fluor-apatite, Laslett et al., 1987) The closure temperature in fission track studies refers to a temperature within the partial annealing zone (PAZ) where more than 50% of the fission tracks are retained. The PAZ is a wide temperature range with gradually decreasing track stability towards higher temperatures. Within the PAZ, tracks are therefore gradually annealed as a function of time and temperature. Track length distributions of tracks which are oriented parallel to the surface and fully enclosed in the crystal (confined tracks, Bhandari et al., 1971) are a diagnostic tool for thermal histories, because shortened tracks indicate longer residence time or higher temperature during this residence in the PAZ than long tracks. This method is herein only applied to apatite as its annealing conditions are better constrained than those of zircon.

For interpretation of the zircon fission track data we apply a closure temperature of 260 ± 30 °C. This temperature is derived from a range of 290–230 °C, which was calculated by Brandon et al. (1998) from experimental data for geologically realistic cooling rates (e.g., between 5 and 500 °C/Ma). It also encompasses values estimated from calibrations using $^{39}$Ar–$^{40}$Ar studies in K-feldspar and biotite (Foster et al., 1996). Eight zircon samples from the Chirripó area have fission track ages between 11.4 ± 1.4 and 8.4 ± 1.2 Ma. Five zircon ages determined on samples from the Buenos Aires area range from 12.4 ± 1.8 to 9.4 ± 1.4 Ma (Table 2). Within 2σ error, individual zircon fission track ages are identical (Figs. 3 and 4).

The fission track ages of seven apatite samples from the Chirripó area range from 4.8 ± 1.6 to 1.7 ± 0.6 Ma. Three apatite samples from the Buenos Aires area gave ages between 4.8 ± 1.6 and 2.5 ± 1.0 Ma (Table 3; Figs. 3 and 4). The closure temperature for fission tracks in apatite is strongly influenced by chlorine concentration. In general, higher chlorine content leads to higher resistance to annealing and a higher closure temperature (Gleadow and Duddy, 1981; Green et al., 1989a,b; O'Sullivan and Parrish, 1995; Kohn and Foster, 1996). Therefore, electron microprobe analysis of the apatite was performed (Gräfe, 1998). Seven apatite samples from the Chirripó area show chlorine concentrations of 1.2–3.5 wt.%. The three apatite samples from the Buenos Aires area have chlorine concentrations of 0.7–1.0 wt.%. Comparing these concentrations with the 0.4 wt.% chlorine content of the Durango apatite standard the apatites from the Chirripó and Buenos Aires study
areas are chlorine-rich apatites. Due to their higher resistance to annealing, the PAZ and the closure temperature may be somewhat higher than the 125–160°C (Green et al., 1989a,b) range observed in Durango apatite. From experimental data (Carlson et al., 1999) from apatites with chlorine concentrations similar to those of the present study, Ketcham et al. (1999) calculated closure temperatures which range from 190 to 150°C. They also calculated the temperature, at which a fission track population fully anneals for a given heating rate, to values ranging from 230 to 170°C (total annealing temperature). Based on the results of their study we apply a closure temperature of 170°C for the high temperature limit of the apatite PAZ (Fig. 6) for the interpretation of the apatite fission track data.

It is important to note that the apparent fission track apatite ages were not corrected for shortened track lengths. Such corrections do not give reliable results, because little is known about the initial track length in chlorine-rich apatite. A correction for shortened tracks would result in ages that are slightly older than the values given in Table 3.

The track length distribution of some samples (Fig. 5) has been investigated in order to determine the time–temperature paths of apatite through the PAZ. The confined track length distributions in samples from both study areas show a large proportion of shortened tracks (Fig. 5). Up to one-third of the confined tracks are between 5 and 10 μm long. For the Chirripó samples the track length distribution diagrams (Fig. 5a–f) show a pronounced maximum of track lengths at values between 14 and 16 μm, indicating rapid final cooling. For the Buenos Aires sample 80 (Fig. 5g), the track length distribution is broader, with the majority of track lengths between 12 and 15 μm, suggesting that cooling through the PAZ was slightly slower than in the Chirripó area.

5. Cooling history of the Cordillera de Talamanca

Knowledge of the time–temperature history of the Cordillera de Talamanca is important for elucidating the active upper plate tectonics during subduction of the Cocos ridge. Geochronological data from this study provides new insight into the response of the upper plate to ridge subduction.
5.1. High-temperature cooling history

The intrusion depth of the Miocene granitoid rocks of the Cordillera de Talamanca was calculated using the aluminum-in-hornblende geobarometer (Johnson and Rutherford, 1989) on sample 48 from the Chirripó area and on samples 78 and 80 from the Buenos Aires area. The results indicate shallow intrusion depths of < 5 km. Emplacement depth can only be estimated since the sample record pressures between 0.4 and 1.2 kb, which fall outside the experimentally calibrated range of 2–6 kb (Johnson and Rutherford, 1989). However, our results are in good agreement with the previously suggested epizonal intrusion of melts at temperatures between 805 and 860 °C (Drummond et al., 1995).

Modeling the temperature evolution for conductive and convective cooling mechanisms for plutons with similar sizes and intrusion depths to those of the Cordillera de Talamanca shows that thermal reequilibration occurs on a time scale of less than 1 Ma and increases the geothermal gradient in the country rock (Norton and Knight, 1977; Harrison and Clarke, 1977).

Stages of cooling are indicated by the Rb–Sr biotite and the zircon fission track ages. The emplacement of the granitoid rocks into the upper crust, subsequent rapid cooling and the similarity of the individual zircon fission track ages, warrants the calculation of a weighted average of the individual zircon fission track ages for each study area. The
The weighted average zircon age for the Chirripó area is 9.7 ± 0.9 Ma (2σ; MSWD = 2.1) and for the Buenos Aires area 10.3 ± 1.2 Ma (2σ; MSWD = 2.0). The Rb–Sr biotite ages of about 10 Ma of the Buenos Aires area and the weighted average zircon ages of about 10 Ma in both areas indicate cooling to at least 260 °C at about 10 Ma and provide a minimum age estimate for emplacement and crystallization of granitoid rocks in the Cordillera de Talamanca. Taking the cooling models for plutons into account, the actual age of emplacement of the granitoid rocks is probably between 11 and 10 Ma.

In the Chirripó area, the weighted mean zircon age of 9.7 ± 0.9 Ma is older than the apparent ⁴⁰Ar–³⁹Ar and the Rb–Sr biotite ages of 8.6–7.5 Ma and the ⁴⁰Ar–³⁹Ar amphibole age of about 8.7 Ma. As the fission track closure temperature of zircon is lower than the commonly accepted closure temperatures for Rb–Sr and K–Ar in amphibole and biotite, this obviously interferes with any straightforward interpretation of the ⁴⁰Ar–³⁹Ar and Rb–Sr ages as cooling ages. Thus, other mechanisms influencing the isotopic signatures of the granitoid rocks need to be considered. Chloritization of biotite and formation of actinolite in the Chirripó samples point to the fact that low-temperature alteration affected the K–Ar and the Rb–Sr systems. Using the geothermometer described by Cathelineau and Nieva (1985) on chlorite–biotite intergrowths in sample 48 indicates formation of chlorite at approximately 227 °C. The possibility of chlorite formation at temperatures from 325 to 155 °C was also documented by Battaglia (1999).
calculations show that alteration can occur at temperatures lower than the closure temperature of fission tracks in zircon.

Clauer (1981) and Clauer et al. (1982) concluded that low-temperature alteration of biotite results in a preferential loss of radiogenic $^{87}\text{Sr}$ relative to Rb, in an

<table>
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<th>Sample</th>
<th>Alt. (m)</th>
<th>Number of grains</th>
<th>$\rho_s \times 10^3$ (cm$^{-2}$)</th>
<th>$\rho_i \times 10^5$ (cm$^{-2}$)</th>
<th>$N_s$</th>
<th>$N_i$</th>
<th>$P_{\chi^2}$ (%)</th>
<th>$U$ (ppm)</th>
<th>$\rho_d \times 10^5$ (cm$^{-2}$)</th>
<th>$N_d$</th>
<th>Age (Ma) $\pm 2\sigma$</th>
<th>Mean track length (μm ± S.D.) (no. of tracks)</th>
<th>Cl (wt.%)</th>
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<td>3.8 ± 1.0</td>
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<td>4.8 ± 1.6</td>
<td>12.8 ± 2.9 (60)</td>
<td>0.7 ± 0.1</td>
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</tbody>
</table>

Alt. — Altitude of sample locality; CI — chlorine concentration (mean) was determined by microprobe analysis (based on 23 oxygen atoms) on 10–15 grains per sample; errors on chlorine concentration are reported as standard deviation of the mean; $\rho_s$ and $\rho_i$ — spontaneous and induced track densities in the apatite sample; $\rho_d$ — density of induced tracks in the dosimeter glass; $N_s$ and $N_i$ — number of the counted spontaneous and induced tracks; $N_d$ — number of counted tracks in the dosimeter glass CN 5; $P_{\chi^2}$ is the probability of obtaining $\chi^2$ values for $v$ degrees of freedom; $U$ — uranium concentration in sample; S.D. — standard deviation of the mean; n.d. — not determined.

The ages are reported as central ages (Galbraith and Laslett, 1985).

Fig. 5. Fission track length distribution diagrams for apatite samples from the Chirripó area (a–f) and from the Buenos Aires area (g); Abbreviations: $n$ — number of tracks counted, $l$ — mean track length, S.D. — standard deviation.

exchange of radiogenic Sr against Sr derived from the environment, and in preferential loss of radiogenic $^{40}$Ar relative to K. Apparent ages then tend to be gradually reset, as a function of the degree of alteration, and have no direct geological significance. The shifts in apparent Rb–Sr and K–Ar ages of biotite caused by alteration are fairly systematic in a sense that the $^{40}$Ar–$^{39}$Ar and Rb–Sr ages are reset to similar degrees. This is exemplified by the similar apparent ages of about 8.6 Ma obtained for both, the $^{40}$Ar–$^{39}$Ar and the Rb–Sr system on biotite sample 48 (Table 1). Therefore, we conclude that the $^{40}$Ar–$^{39}$Ar and Rb–Sr data from the Chirripó area represent disturbed ages due to partial resetting by hydrothermal alteration. With our data it is impossible to precisely date the alteration, which may have occurred as a prolonged process. Assuming that resetting of the biotite ages was partial, hydrothermal alteration must have continued to times younger than about 7.5 Ma, which is the youngest biotite age in the Chirripó area, since apparent ages tend to become younger during alteration. In the Buenos Aires area, the three identical Rb–Sr biotite ages and the weighted mean fission track zircon age indicate that alteration did not affect the samples from this area.

5.2. Low-temperature cooling history

The low-temperature cooling history is derived from apatite fission track thermochronology. Apart from one group of related papers, there is little published of experimentally determined annealing data for apatite with as high chlorine content as documented herein (Carlson et al., 1999; Donelick et al., 1999; Ketcham et al., 1999). This precludes rigorous modeling of the sample cooling histories. The closure temperature for the high-chlorine apatites analyzed in this study is estimated to $170 \pm 20 ^{\circ}C$ (see discussion above). The large proportion of shortened tracks in both sampling areas indicate either prolonged residence of the samples in the PAZ, or an episode of reheating of the apatites. In the case of a prolonged residence in the PAZ (Fig. 6, model path a), the samples probably resided there up to one third of the time since the samples entered the PAZ, since shortened tracks make up about one third of the total number of confined tracks. This suggests that the samples entered the PAZ at about 5 Ma (Fig. 6). If the samples experienced reheating (Fig. 6, model path b), the thermal pulse could have been due to adakite magmatism beginning at about 3.6 Ma (de Boer et al., 1995). With the current data we cannot distinguish between these two scenarios. However, in any case, the samples resided for significant times at temperatures below the high temperature limit of the PAZ before final cooling through the apatite fission track closure temperature.

Most importantly, both apatite ages and track length distributions of the samples are similar, regardless of topographic elevation of sample localities (Figs. 3–5). This observation and the overlap of age error intervals justify the use of weighted average apatite ages instead of individual apatite ages. The weighted average of samples from the Chirripó area is $3.1 \pm 1.4$ Ma ($2\sigma$; MSWD = 1.5) and for those from the Buenos Aires area $3.0 \pm 1.4$ Ma ($2\sigma$; MSWD = 0.8). The similar ages and track length distributions indicate that all samples experienced a similar thermal history. The abundance of shortened tracks point to a prolonged residence time in the PAZ or reheating of the samples. Rapid exhumation occurred thereafter, as indicated by the large amount of long tracks. Slow exhumation can be ruled out as it would have resulted in correlations between topography, age and track length distribution. Rapid exhumation took place at about 3 Ma, as indicated by the weighted mean apatite ages.

With our data, we cannot establish the mechanisms for exposing samples with similar ages and thermal histories at different altitudes. A possible cause for the lack of correlation between apatite fission track ages and topography can be that during fast exhumation the isotherms became more parallel to topography (Stüwe et al., 1994; Mancktelow and Grasemann, 1997). Differential displacement within the granitoids may account for the missing correlation, too. Fault kinematic, seismic and geodetic data from the northern part of the Cordillera de Talamanca show that NW striking faults may accommodate uplift and exposure within the northern Cordillera de Talamanca (Marshall et al., 2000). The authors also suggest that in response to shallow subduction of the Cocos ridge, deformation propagated from the fore-arc area into the magmatic arc. In contrast to the northern part of the Cordillera de Talamanca, tectonic structures in the central part cannot be mapped due to dense vegetation. However, it is likely that deformation caused by the horizontal shortening due to increased basal traction from the shal-
lowly subducting Cocos ridge is not only restricted to the northern part of the Cordillera de Talamanca, but similarly affects the central part. Isostatic reequilibration of the buoyant shallowly subducting Cocos ridge probably contributed to the exhumation of the cordillera. Rapid exhumation at about 3 Ma accounts very well for the geochronological data obtained in this study, such as (1) the lack of a fission track apatite age–elevation correlation; (2) the identical weighted mean fission track apatite ages of 3 Ma in different study areas; (3) the similar fission track length distribution in samples with up to 2500-m difference in elevation.

Considering the geologic setting in southern Costa Rica we attribute this exhumation to the arrival of the buoyant Cocos ridge underneath the Cordillera de Talamanca which probably caused isostatic reequilibration and subhorizontal shortening.

6. Tectonic model and conclusions

The significance of the Cocos ridge subduction to the upper plate geodynamic evolution is delineated in Fig. 7. During the Late Miocene, cold, dense oceanic crust was subducted below the western margin of the Caribbean plate (Donnelly, 1985), accompanied by the formation of the magmatic arc (Fig. 7A). Zircon fission track and Rb–Sr biotite ages combined with geobarometry suggest shallow emplacement of granitoid melts in the Cordillera de Talamanca between 11 and 10 Ma. At this time, the subduction angle was probably about 60°, as currently observed below central Costa Rica (Protti et al., 1995).

The geochronological data obtained in this study allows the calculation of the onset of ridge subduction...
using the following parameters: (1) the final cooling event at about 3 Ma as suggested by apatite fission track ages; (2) the distance of ca. 130 km between the Cordillera de Talamanca and the present-day trench; (3) the present-day convergence rate between the Cocos and Caribbean plates (9.6 cm/year; DeMets et al., 1990; Seno et al., 1993); and (4) the current subduction angle of 30° (Protti et al., 1995). Calculations using these parameters show that the Cocos ridge was located below the Cordillera de Talamanca about 1.5 Ma after it entered the trench (Fig. 7B). Considering the uncertainties on the subduction geometry parameters and the age of final cooling of the rocks, the uncertainty on the beginning of Cocos ridge subduction is calculated to be about 1 Ma, suggesting that ridge subduction began between 5.5 and 3.5 Ma.

For the Early Pliocene the following scenario is proposed: At about 3 Ma, ca. 150 km of the Cocos ridge was subducted at a low-angle, replacing the former asthenospheric wedge underneath the Cordillera de Talamanca (Fig. 7B). Subsequent exhumation of the granitoid rocks of the Cordillera de Talamanca occurred by horizontal shortening and isostatic reequilibration.

The results from this study on the onset of Cocos ridge subduction correlates well with the results from other studies such as the paleobathymetric uplift ages of 3.6 and 1.6 Ma for the Burica Peninsula and the Limón basin, respectively (Collins et al., 1995) and the development of thrust faults in the fore-arc basins attributed to increased coupling of the upper and lower plate starting at about 5 Ma (Kolarsky et al., 1995). The cessation of calc-alkaline magmatism has been explained by the shallowing of the subduction angle (Drummond et al., 1995). Adakite volcanism starting at 3.6 Ma was attributed to the presence of the Cocos ridge underneath the Cordillera de Talamanca, (Defant and Drummond, 1990; de Boer et al., 1995; Abratis and Wörner, 2001) supporting the results of our study.

Our study on the time–temperature evolution and the exhumation history of the Cordillera de Talamanca provides a minimum emplacement age for granitoid rocks in the Cordillera de Talamanca of about 10 Ma based on weighted mean fission track zircon ages of $9.7 \pm 0.9$ Ma for the Chirripó area and of $10.3 \pm 1.2$ Ma for the Buenos Aires area, and Rb–Sr biotite ages of 10 Ma for the Buenos Aires area. The onset of Cocos ridge subduction between 5.5 and 3.5 Ma can be inferred from the apatite fission track weighted mean ages, which are consistent with independent
evidence from the tectonic, magmatic and volcanologic record. The final cooling of the upper crustal rocks is mainly governed by isostatic rebound and tectonic processes due to the shallowly subducting Cocos ridge.

Acknowledgements

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Appendix A. Analytical methods

A.1. 40Ar–39Ar dating

Amphiboles and biotites were separated using standard mineral separation techniques. From each amphibole sample, 40–80 mg was handpicked to achieve visual purity. Biotite was purified by grinding in a mortar with alcohol. Then, 30–40 mg of each biotite sample was then handpicked to achieve visual purity. The samples were step-heated in a double vacuum resistance oven connected to a MAP 215-50B mass spectrometer. The isotope data were corrected for mass spectrometer background, mass fractionation, and 37Ar decay. The ages are corrected for interfering isotopes, as described in Belluso et al. (2000).

A.2. Rb–Sr dating

Rb and Sr concentrations of whole rock samples were determined by X-ray fluorescence analysis, those of the biotite samples by the isotope dilution method. For biotite and whole rock isotope analysis, 40–50 mg of each sample was dissolved in HF–HClO4 in PFA vessels by open-beaker dissolution. The second step of sample decomposition was done in Teflon bombs at 180 °C for 1 week, to ensure decomposition of refractory phases such as zircon. Apatite samples were dissolved in 2 N HNO3. Biotite samples were spiked with a 84Sr–87Rb isotope tracer solution before decomposition. Four analyses of the NBS-987 Sr-standard gave an average 87Sr/86Sr value of 0.710244 ± 0.000010. The isotopic compositions were measured on a Finnigan MAT 262 mass spectrometer at the University of Tübingen. Sr was analyzed using a single W filament and a Ta activator solution. Isotope analysis was done using a dynamic multi-cup measuring routine which monitors for interfering Rb. 87Sr/86Sr was corrected for instrumental mass fractionation with a 86Sr/88Sr ratio of 0.1194. 87Rb/85Rb ratios were determined on a Re-double-filament configuration and using a static data collection procedure. A fractionation factor of 0.3% per atomic mass unit was applied to correct for mass fractionation. For a more detailed description of the analytical methods used, see Hegner et al. (1995).

A.3. Fission track dating

Apatite and zircon fission track ages were determined using the external detector method. Apatite grains were embedded in epoxy, then polished, and etched with 1% HNO3 for 210 s at room temperature. Zircon grains were embedded in Teflon, then polished, and etched in a KOH–LiOH eutectic melt at 205 °C for periods between 40 and 87 h. The mineral mounts were covered with low-uranium muscovite detectors and irradiated together with neutron dose monitor glasses (CN 2 for zircon, and CN 5 for apatite). Samples were irradiated at the Risø reactor in Roskilde, Denmark. The thermal neutron doses were determined from the track densities in muscovite detectors that covered the glass standards during irradiation. Microscopic analysis was done under oil immersion at 1250 x magnification. Ages were calculated using the zeta calibration method (Hurford and Green, 1983). All ages are reported as central ages (Galbraith and Laslett, 1985). Zeta values of 354 ± 9 for apatite and of 151 ± 3 for zircon were used for the age calculations. The following age standards were used for the determination of the zeta values: Fish Canyon Tuff (apatite and zircon), Durango (apatite), and Tardree Rhyolite (zircon). In total, 13 apatite and zircon samples were analyzed. Data from four out of
13 apatite samples were discarded because of a very high number of dislocations in the crystal lattice, which strongly interfered with the proper identification of spontaneous fission tracks. Lengths of confined tracks could only be measured on seven out of 13 apatite samples, despite examining up to eight mounts per apatite sample for confined track measurements. The apatite track length distribution measurements were carried out on sub-horizontal tracks wholly included in the host mineral using faces parallel to the c-axis of the mineral. Determination of confined fission track lengths on the other six samples was not possible because either (1) the track density was too low, (2) the dislocation density was too high, or (3) there was not enough material to permit the measurement of a representative number of confined tracks.

References


